The westwards propagation of the opening of the Mesozoic Tethys and the breack-up of Pangea.

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Thematic Article

Tethyan ophiolites and Pangea break-up

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Abstract The break-up of Pangea began during the Triassic and was preceded by a generalized Permo-Triassic formation of continental rifts along the future margins between Africa and Europe, between Africa and North America, and between North and South America. During the Middle–Late Triassic, an ocean basin cutting the eastern equatorial portion of the Pangea opened as a prograding branch of the Paleotethys or as a new ocean (the Eastern Tethys); westwards, continental rift basins developed. The Western Tethys and Central Atlantic began to open only during the Middle Jurassic. The timing of the break-up can be hypothesized from data from the oceanic remnants of the peri-Mediterranean and peri-Caribbean regions (the Mesozoic ophiolites) and from the Atlantic ocean crust. In the Eastern Tethys, Middle–Late Triassic mid-oceanic ridge basalt (MORB) ophiolites, Middle–Upper Jurassic MORB, island arc tholeiite (IAT) supra-subduction ophiolites and Middle–Upper Jurassic metamorphic soles occur, suggesting that the ocean drifting was active from the Triassic to the Middle Jurassic. The compressive phases, as early as during the Middle Jurassic, were when the drifting was still active and caused the ocean closure at the Jurassic–Cretaceous boundary and, successively, the formation of the orogenic belts. The present scattering of the ophiolites is a consequence of the orogenesis: once the tectonic disturbances are removed, the Eastern Tethys ophiolites constitute a single alignment. In the Western Tethys only Middle–Upper Jurassic MORB ophiolites are present – this was the drifting time. The closure began during the Late Cretaceous and was completed during the Eocene. Along the area linking the Western Tethys to the Central Atlantic, the break-up was realized through left lateral wrench movements. In the Central Atlantic – the link between the Western Tethys and the Caribbean Tethys – the drifting began at the same time and is still continuing. The Caribbean Tethys opened probably during the Late Jurassic–Early Cretaceous. The general picture rising from the previous data suggest a Pangea break-up rejuvenating from east to west, from the Middle–Late Triassic to the Late Jurassic–Early Cretaceous.

Key words: Caribbean Tethys, Eastern Tethys, Jurassic, ophiolites, Paleozoic Pangea break-up, Triassic, Western Tethys.

INTRODUCTION

During Late Permian to Early Triassic time, the eastern border of Pangea was the peri-equatorial Paleotethys Ocean, the westernmost extension of the Panthalassa Ocean, which developed during the Paleozoic.

The break-up of Pangea and the consequent formation of two supercontinents (Gondwana to the south and Laurasia to the north) started during the Permian and continued throughout the Middle Triassic to Late Jurassic. The break-up is witnessed by an east–west-trending equatorial oceanic basin (or several basins according to many authors), which, during the Middle Triassic,

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crossed the eastern portion of the equatorial Pangea. It is still a matter of debate whether this oceanic basin was a new, developing branch of the Paleotethys Ocean. Alternatively, the basin could have formed as an incipient opening of a new ocean, the Eastern Mesozoic Tethys or 'Neotethys', which, during the Triassic, began to separate Gondwana (Adria Promontory) from Laurasia. The location of the western termination of this oceanic basin is uncertain, although several lines of evidence indicate that it can be located in the region now occupied by the Western Carpathians. What is more certain is that westward, along the paleo-equatorial zone, the future boundary between Gondwana and Laurasia was occupied by the orogenically mature and partially eroded Hercynian chain (Fig. 1).

During the Middle Triassic to Late Jurassic, this orogenic area of Pangea was subjected to widespread and intense extensional tectonics, probably as a result of transtension along a large left-lateral east–west shear zone along the future boundary between Laurasia and Gondwana (Fig. 2). This area of thinned continental crust was covered by shallow seas, and with progressive crustal thinning it evolved into a series of platforms and basins, trending approximately east–west. Some of the basins later evolved to the oceanic stage, as a result of increasing left-lateral strike-slip tectonics within Pangea. Along the western side of Pangea, rifting dominated throughout Triassic and Early Jurassic time. These rift basins indicate the incipient separation between Africa and Eurasia to the east and the Americas to the west through the opening of the Central Atlantic and, successively, of the North Atlantic Ocean. A tectonic line extending east-northeast from Gibraltar was the forerunner to the separation of Africa and Adria from Europe.

Rifting phases were also present to the north of the Gulf of Mexico, although the later separation between North and South America occurred in a more southern zone, in the Caribbeans.

A number of paleogeographic reconstructions for the break-up of Pangea during the Triassic to Late Jurassic have been proposed for different regions:

1. peri-Mediterranean (e.g. Dewey *et al.* 1973; Smith *et al.* 1973; Laubscher & Bernoulli 1977; Channell *et al.* 1979; Abbate *et al.* 1980, 1986; Robertson & Dixon 1984; Sengor 1984; Dercourt *et al.* 1986, 1993; Gealey 1988; Dal Piaz *et al.* 1995; Robertson *et al.* 1996; Stampfli & Mosar 1999; Stampfli *et al.* 2002; Golonka 2004)
2. peri-Central Atlantic Ocean (e.g. Dewey *et al.* 1973; Smith *et al.* 1973; Michard 1976; Laubscher & Bernoulli 1977; Smith & Briden 1977; Channell *et al.* 1979; Klitgord & Schouten 1986; Smith & Livermore 1991; Dercourt *et al.* 1993; Channell & Kozur 1997)
3. peri-Caribbean (e.g. Pindell & Dewey 1982; Gose 1983; Duncan & Hargraves 1984; Pindell

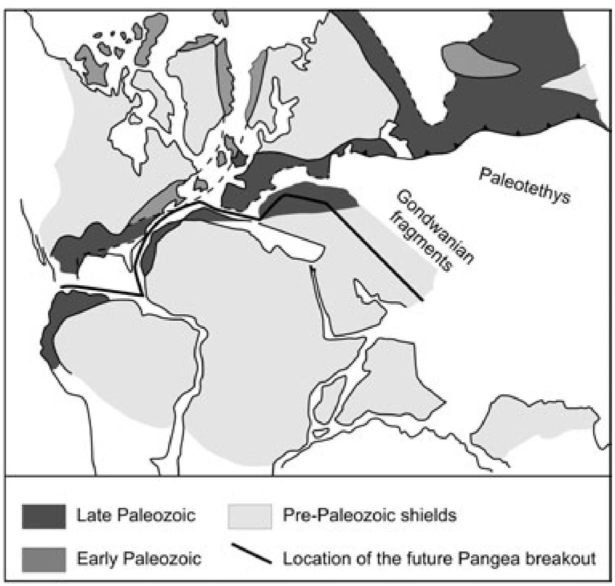

Fig. 1 The Pangea reconstruction showing the Paleozoic orogenic belts (redrawn from Smith & Briden 1977).

Fig. 2 Reconstruction of Pangea during Middle–Late Triassic, the dark gray areas are the rift basin systems, mainly located along the future break-up of the supercontinent (after Stampfli & Borel 2002; slightly modified).

1985, 1994; Klitgord & Schouten 1986; Ross & Scotese 1988; Pindell & Barrett 1990; Stephan *et al.* 1990; Bartok 1993; Giunta 1993; Meschede 1994; Montgomery *et al.* 1994a; Beccaluva *et al.* 1996; Giunta *et al.* 1998; Meschede & Frisch 1998; Pindell *et al.* 1998; Kerr *et al.* 1999).

In the present paper, the distribution of the Triassic continental rift basins along equatorial Pangea, and the location of the Triassic, Jurassic and Cretaceous ophiolite belts (i.e. oceanic areas) in the peri-Mediterranean and peri-Caribbean orogenic domains and in the Central Atlantic are discussed. A timeline for the break-up of Pangea, a discussion on the mechanisms that controlled it and a reconstruction of the paleogeodynamic domains involved in this global event will be presented. The reconstructions are based mainly on a review of a large amount of stratigraphic and petrologic evidence collected in the region during the past few decades. In particular, the chronological reconstructions are mostly based on radiolarian biostratigraphy data, because of the scarcity of radiometric dates. Finally, an updated reconstruction of the Triassic–Jurassic Tethyan basins of the peri-Mediterranean area during the separation of Laurasia and Gondwana will be proposed.

For the reconstructions in the present paper, the Geologic Time Scale of the International Commission on Stratigraphy has been chosen (Gradstein *et al.* 2004). There is a problem concerning the reliability of K–Ar dating as suggested by comparison with the results obtained by the Ar–Ar method on the same ophiolite units (e.g. the ferrodiorites of La Bartolina, Italy, gave 157 Ma with Ar–Ar, and approximately 200 Ma with K–Ar; Bortolotti *et al.* 1995). Similar problems were also found when radiometric dating was compared to radiolarian biostratigraphy (e.g. in an ophiolite section in the Apuseni Mountains, the radiometric date of pillow basalts was younger than the biostratigraphic age of the overlying radiolarian cherts).

TRIASSIC PELAGIC BASINS NEAR FUTURE (JURASSIC) CONTINENTAL MARGINS

Evidence of Triassic rifting is common in many orogenic areas of the world, especially along the orogenic belts, which in later times will become ocean basins. The focus now will be on the evolution of rift basins as documented in three regions: the peri-Mediterranean, the Central Atlantic and the Gulf of Mexico–Caribbean.

PERI-MEDITERRANEAN REGION

The continental margins of the peri-Mediterranean region developed during the Middle–Late Triassic, although their subsidence started as early as the Permian. However, the widespread marine invasion of these regions occurred mainly during the Middle Triassic (Ciarapica *et al.* 1986; Zappaterra 1990; Ciarapica & Passeri 2002; Bortolotti *et al.* 2004a). Since the Early Triassic, the areas of the future location of the continental margins were already fragmented, forming an articulated system of platforms and basins (Figs 3,4). However, not all areas affected by pre-Carnian extensional tectonics developed into continental margins. Many of the pre-Carnian basins also persisted throughout the Triassic (e.g. the Lagonegro and Pindos Basins). Pre-Carnian basins occur in the Alpine–Apenninic and Dinaric–Hellenic–Pontic chains (the future Apulia–Adria–Menderes Plate) parallel or *en echelon* to each other, with respect to the future oceanic margins.

The formation of intracontinental basins and platforms during Triassic time was the result of the incipient crustal thinning of Pangea. Moreover, several lines of evidence suggest that strike-slip tectonics were widespread since the Permian and were actively forming more or less restricted pull-apart basins (see Vai 1991).

During the Anisian–Ladinian, several of the basins of the peri-Adriatic region were affected by volcanism, from acidic rocks to alkaline-basalts (e.g. Zappaterra 1990; references therein; Fig. 5).

Despite the tectonic deformation and the translational and rotational movements experienced during later orogenesis, the relative locations and alignments of the platform and the basins, of the volcanics, and of the Mesozoic ophiolitic complexes can be considered to be not far from the original relationships with the old continental–oceanic margins, which were at that time all roughly aligned east–west (Ciarapica *et al.* 1986).

CENTRAL ATLANTIC

As shown in Figure 6, in the area where the Central Atlantic developed, several basins elongated parallel to the future continental margins of Africa and North America developed during the Triassic (Van Houten 1977; Manspeizer *et al.* 1978; Emery & Uchupi 1984; Swanson 1986; Manspeizer 1988; Smith & Livermore 1991). Coeval rift magmatism was also developing along both the American and the African margins bordering the Central Atlan-

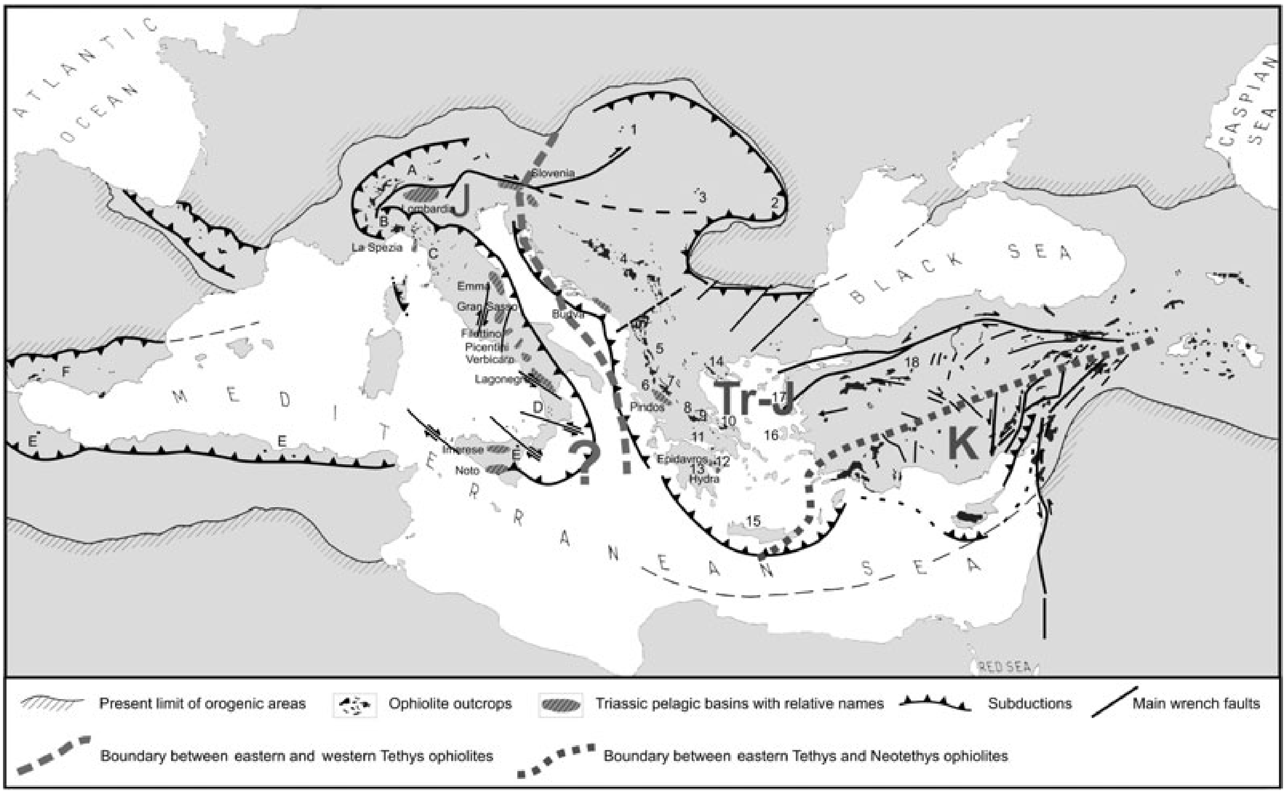

Fig. 3 The circum-Mediterranean Upper Triassic main rift basins and the ophiolite outcrops subdivided in three areas. A, Central Alps; B, Western Alps; C, Northern Apennines; D, Southern Apennines; E, Sicily and North Africa; F, Betic Cordillera; J, Jurassic ophiolites of the Western Tethys; K, Cretaceous ophiolites of the Neotethys; Tr-J, Triassic and Jurassic ophiolites of the Eastern Tethys. Tr-J: 1, Western Carpathians; 2, Dobrogea; 3, Apuseni Mountains; 4, Dinarides; 5, Mirdita; 6, Pindos; 7, Vourinos; 8, Koziakas; 9, Othrys; 10, Euboea; 11, Kerassies; 12, Argolis; 13, Peloponnesos; 14, Guevgueli–Chalkidiki; 15, Crete; 16, Chios; 17, Lesvos; 18, Northern and Central Turkey.

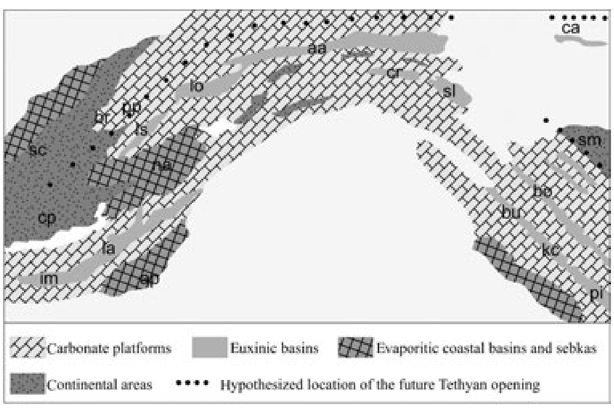

Fig. 4 Schematic paleogeography of the peri-Adriatic region during Norian time (redrawn from Ciarapica *et al.* 1986). aa, Austro–Alpine; ap, Apulia; bo, Bosnia; br, Briançonnais Ridge; bu, Budva; ca, Carpathians; cp, Calabria–Peloritani Massif; cr, Carnian; im, Imerese; kc, Krasta–Cukali; la, Lagonegro; lo, Lombardo; ls, La Spezia; na, Northern Apennines; pi, Pindos; sc, Sardinia and Corsica; sl, Slovenian.

tic Ocean (Van Houten 1977; Manspeizer *et al.* 1978; Manspeizer 1988). Different types of Middle Triassic–Early Jurassic rifting magmatism and their distribution in the Atlantic Morocco margin are shown in Figure 7.

PERI-CARIBBEAN ZONE

In the peri-Caribbean zone, the paleogeography of the Triassic continental margins is not so well documented. However, they are well described in adjacent areas, such as along the northern border of the Gulf of Mexico (Thomas 1988). According to Thomas (1988), this zone is characterized by the development of Lower–Middle Mesozoic basins and associated magmatisms, of often uncertain age, but generally ranging from Triassic (linked with the Red Beds of the Eagle Mills Formation; Vernon 1971) to Late Cretaceous(?) (73 ± 2.9 Ma, K–Ar; Sundeen & Cook 1977).

In the Alabama–Arkansas Fault system (Western Mississippi), undated magmatic rocks lie below Jurassic sediments. The basins are very far from the peri-Caribbean zone, but they could be related to a Tethysian basin represented now by the deep central portion of the Gulf of Mexico, as suggested by Ross and Scotese (1988).

According to Bartok (1993), Triassic rifting also occurred in the northern margin of South

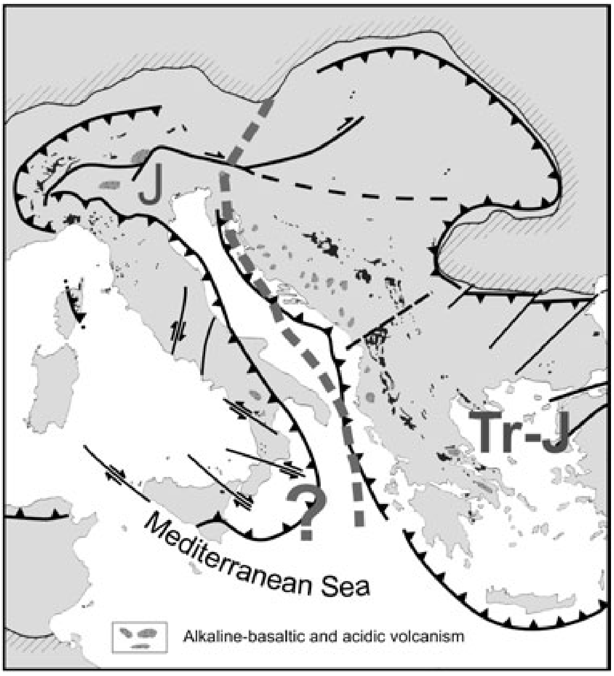

Fig. 5 The circum-Mediterranean Triassic alkaline–basaltic and acidic volcanism. For other symbols, see Figure 3.

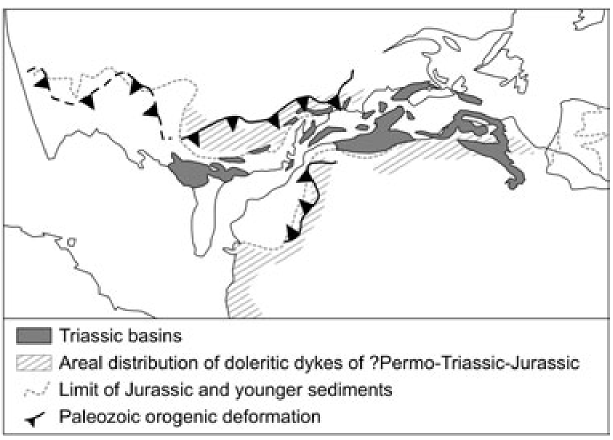

Fig. 6 Peri-Atlantic Triassic basins and areal distribution of the (Permian)–Triassic–Jurassic doleritic dykes (from Michard 1976; modified).

America. The rifting in the two margins occurred in two stages: first during the Triassic (Anisian–Ladinian) and then during the Jurassic (Bajocian–Bathonian). The first phase occurred in a northern location, parallel to the southern North America Paleozoic orogenic belts. The second phase developed more to the south, between the Maya Block and the South America continent, and parallel to the Upper Paleozoic orogenic belts.

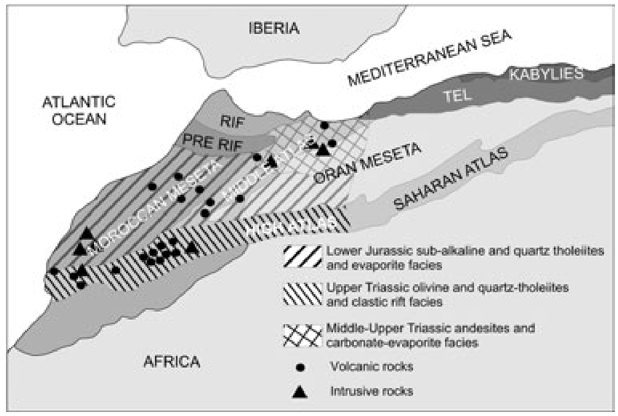

Fig. 7 Distribution of the Middle Triassic–Lower Jurassic rifting magmatism in the Atlantic Morocco margin (from Manspeizer *et al.* 1978; redrawn).

THE SEA FLOOR SPREADING STAGE

DISTRIBUTION AND AGE OF THE OPHIOLITES IN THE PERI-MEDITERRANEAN AND PERI-CARIBBEAN DOMAINS

In the peri-Mediterranean region, the Lower–Middle Mesozoic oceanic areas are classically divided into three groups (Ohnenstetter *et al.* 1979): (i) the Western Carpathian ophiolites (Pieniny–Meliata–Dinaric and Hellenic–Pontic Belts) pertain to the Eastern Tethys Ocean and formed during the Middle Triassic–Middle Jurassic; (ii) the ophiolites of the Betic Cordillera–Apenninic–Alpine Belts pertain to the Western Tethys Ocean and are Middle–Late Jurassic in age; and (iii) the ophiolites of the Taurus Belt pertain to the Neotethys Ocean and formed during the Late Cretaceous. In Figure 3, all the cited ophiolitic zones are located.

The Eastern Tethyan ophiolites did not suffer orogenic metamorphism (except for the ophiolites of the Pieniny Mountains, the Western Carpathians, the Rhodope Massif and east of the Vardar Zone, and some ophiolite units of the Pontic Ranges). By contrast, metamorphic, slightly metamorphic and non-metamorphic ophiolite units of the Western Tethys are well documented. The metamorphic ophiolites (from greenschist to blueschist facies) are present mainly in the Alps, in Corsica, in the Southern Apennines and in the Betic Cordillera. The non-metamorphic ophiolites occur mainly in the Northern Apennines and, subordinately, in the Alps, in Corsica and in the Southern Apennines.

The ophiolites cropping out in the peri-Caribbean belts are commonly considered to belong to the Caribbean Tethys. They are located at the margin of the Caribbean Plate, and their age ranges between Jurassic and (?Early) Cretaceous. It is worth nothing that Cretaceous volcanic arc rocks occur close to the outcrops of ophiolite. The ophiolitic sequences located in the outer border of the Caribbean Plate are mainly metamorphic; however, non-metamorphic ophiolites are also present in Cuba, Puerto Rico and in the Lesser Antilles. The inner portion of the Caribbean Plate consists of a basaltic plateau of Cretaceous age.

Between the two Tethyan domains, the Central Atlantic lithosphere should be considered, in which the older ages are referable to the Middle Jurassic.

EASTERN TETHYS

Triassic–Jurassic ophiolites

In the Eastern Tethyan domain (Fig. 3), together with the more widespread Middle and Upper Jurassic mid-oceanic ridge basalt (MORB) and island arc tholeiite (IAT) ophiolites, Triassic MORB ophiolites are also well documented in the Migdalitsa Ophiolitic Complex of the Argolis Peninsula, in southern Greece (Bortolotti *et al.* 2003, 2004c) and in the Rubik Complex of Albania (Bortolotti *et al.* 1996, 2004b,c). In some other localities in Greece and northern Turkey (e.g. Pindos, Jones & Robertson 1991; Euboea, Danelian & Robertson 2001; Othrys and Koziakas, Saccani *et al.* 2003) MOR basalts associated, but not clearly linked, with Triassic cherts are also described.

In the Western Carpathians, Dinarides and Pontides, a widespread mélange of Cretaceous age contains blocks of ophiolitic rocks, mainly basalts with scattered chert levels or large blocks. The cherts yielded radiolarian assemblages indicating both Jurassic and Triassic ages. Unfortunately, at present, no clear stratigraphic relationships between Triassic radiolarian sequences and basalts of MORB affinity have been found.

In Dobrogea (Romania), Lower–Middle Triassic basalts derived from a MORB-type asthenospheric mantle and compatible with a genesis in a rifting setting have been reported (Saccani *et al.* 2004b). In the Mamonia mélange (Cyprus), Triassic MOR and/or within-plate basalts (WPB) have been observed (Dilek & Flower 2003). These could be interpreted as a product of magmatism in deep basins on a thinned continental crust and not as remnants of a Triassic ocean floor sited south of the Menderes Massif.

The major and better known ophiolitic areas are reviewed below (see Fig. 3). Figure 8 displays all the chronological data found in the literature.

Western Carpathians

In the Pieniny Belt, a Cretaceous flysch includes ophiolite pebbles that have an oceanic origin and a

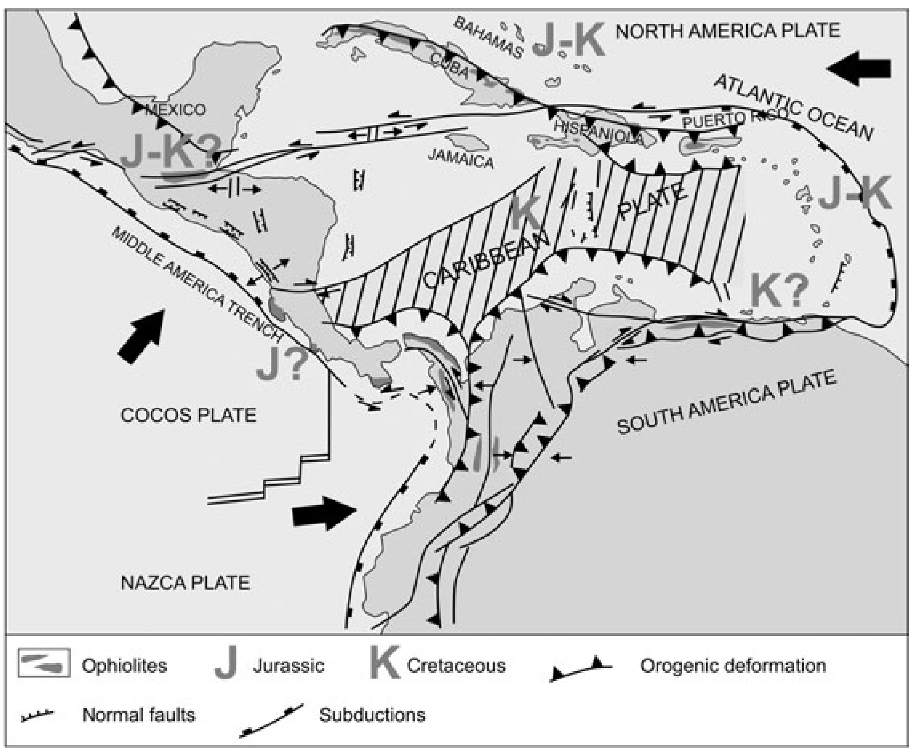

Fig. 8 Distribution of the ophiolites in the peri-Caribbean area.

blueschist facies of metamorphism. In a tholeiitic basalt, a glaucophane yielded an Ar–Ar date of 155.4 ± 0.6 Ma (Oxfordian–Kimmeridgian boundary; Dal Piaz *et al*. 1995). The Pieniny constitute the northern termination of the Vardar Ocean, which began to close at the Oxfordian–Kimmeridgian boundary (Dal Piaz *et al*. 1995).

The Meliata unit contains blocks of ophiolitic rocks (serpentinites and gabbros) and pillow lavas. The latter include thin intercalations of cherts that yielded Ladinian to early Carnian ages (Kozur 1991). Unfortunately, in Kozur's study the geochemistry of the pillow lavas is not reported. However, in the Bodva Valley ophiolite complex, part of the Meliata unit, the MORB character was confirmed (see Horvath 1997). Ar–Ar dates of 150–165 Ma (Bathonian–Kimmeridgian) have been provided by Maluski *et al.* (1993) and Faryad and Henjes-Kunst (1997) for the blueschist metamorphism of this unit.

In the Bükk Mountains, upper Aalenian–Bajocian radiolarian assemblages were found in chert levels linked to the pillow lavas (Kozur 1991). These basalts are part of the Bükk Mountains ophiolite, and show a geochemical imprint of ocean-floor tholeiites (Balla & Dobretsov 1984).

West of the Bükk Mountains, in the Darnó Hill, a section of radiolarian cherts, probably linked to nearby weathered basalts, yielded an early Carnian age (De Wever 1984; Kozur 1991; Dosztály & Józsa 1992). The gabbros of the Darnó Hill yielded K–Ar dates of 152, 166 and 175 Ma (Middle Jurassic; Árva-Sós & Józsa 1992).

Dobrogea

In north Dobrogea (Macin and Niculitel areas) Triassic pillow lavas of MORB-type mantle probably represent the product of basic magmatism in an extensional tectonic setting, possibly an aborted rift, linked to the opening phase of the Triassic Tethys (Seghedi 2001; Saccani *et al*. 2004b).

Apuseni Mountains

The Apuseni Mountains, located in the hinterland of the Southern Carpathians, a Middle Jurassic ophiolitic sequence (Savu *et al*. 1981, 1994; Bortolotti *et al*. 2002, 2004d) covered by Upper Jurassic calc-alkaline volcanics (Savu *et al*. 1981; Bortolotti *et al*. 2002, 2004d) crops out. The ophiolitic sequence consists of scarce ultramafic cumulates, gabbros and basalts, of MORB-type (Savu *et al.* 1981, 1994; Saccani *et al*. 2001). A thin level of cherts yielded a radiolarian assemblage dated Oxfordian (Lupu *et al*. 1995). K–Ar data from the ophiolitic basalts range from 138.9 ± 6.0 to 167.8 ± 5.0 Ma (Middle–Late Jurassic; Nicolae *et al.* 1992). The Upper Jurassic–Lower Cretaceous calc-alkaline volcanics at the top of the sequence are interpreted as being formed in an island arc setting, and indicate the development of a subduction zone at the Jurassic–Cretaceous boundary.

Dinarides

In Croatia, northeast of Zagreb in the Mount Kalnik and Medvednica areas, an ophiolitic mélange crops out. The mélange contains fragments and large blocks of ultramafic rocks, gabbros and basalts, in some places interbedded with cherts. These cherts yielded radiolarian assemblages of late Ladinian to late Carnian age at the Medvednica and middle Carnian to Norian age at the Kalnik Mount (Halamić & Goričan 1995). Two K–Ar dates on gabbros and basalts give 185.0 ± 6.0 Ma and 189.0 ± 6.7 Ma (Pliensbachian), respectively. The basic rocks display tholeiitic affinity (Pamić 1997).

In the central Dinarides (Pamić *et al*. 2002), an ophiolitic nappe (Dinaride ophiolite zone nappe) comprises an ophiolitic mélange including ophiolitic rocks and fragments of their cover. However, no reliable age data from these rocks are yet available. The mélange extends southwards and can be correlated with the Mirdita Rubik Complex in Albania. The upper portion of the mélange consists of large blocks of ophiolitic rocks, with ultramafites (prevailingly lherzolites in the western outcrops and harzburgites to the east), cumulate rocks and basalts, sometimes linked to chert levels. In general, the nappe is unconformably overlaid by Upper Jurassic–Upper Cretaceous clastic sequences. A lateritic level covered by shallow water sediments located at the top of the Zlatibor Ophiolitic Massif yielded an Oxfordian age (Bortolotti *et al*. 1971).

In some localities, the base of the ophiolitic masses is characterized by metamorphic soles. Few radiometric K–Ar dates of the metamorphic soles are available: (i) at the base of the Zlatibor Massif, from 175.0 ± 11.0 to 172.0 ± 8.0 Ma (Toarcian–Aalenian; Spray *et al.* 1984); (ii) at the base of the Krivajah–Koniuk Massif 161.0 ± 4.0 Ma (Callovian–Oxfordian boundary; Spray *et al.* 1984); (iii) at the base of the Brezovica Massif, from 179.0 ± 6.0 to 159.0 ± 5.0 Ma (Toarcian–Oxfordian; Spray *et al.* 1984) and from 169.0 ± 3.0 to 161.5 ± 1.2 Ma (Bajocian–Callovian; Okrusch *et al.*

1978); and (iv) at the base of the Banovina ophiolite 166.0 ± 10.0 and 160.0 ± 10.0 (Bathonian–Callovian; Majer & Lugović 1985 in Pamić *et al.* 2002).

Mirdita (Albania)

The Mirdita Ophiolitic Nappe is located at the top of the Adria units. This nappe is composed of two main tectonic units: the Rubik complex and the overlying ophiolite unit (Bortolotti *et al.* 1996, 2004c; references therein).

A thick layer of MOR basalts with thin intercalations of radiolarian cherts generally lies near the top of the Rubik complex. The radiolarian cherts yielded radiolarian assemblages of Middle and Late Triassic age (Marcucci *et al.* 1994; Chiari *et al.* 1996; Bortolotti *et al.* 2004b).

The ophiolitic unit can be subdivided into two subunits: (i) a western belt with a sequence of serpentinized lherzolites, a gabbroic complex and basalts with MORB affinity – in the upper part of the sequence, intercalations of IAT and MOR-IAT basalts, and boninitic dykes also occur; and (ii) an eastern belt with a sequence constituted by serpentinized harzburgites, an intrusive sequence and volcanics (basalts–andesites–dacites and rhyolites) indicating a supra-subduction zone (SSZ) setting (Beccaluva *et al.* 1994; Bortolotti *et al.* 1996, 2004c; references therein).

Several Ar–Ar dates are available. In the eastern belt a phlogopite vein cutting the harzburgite foliation yielded dates of 159.7 ± 2.6 and 163.9 ± 3.9 Ma (Callovian–Oxfordian), and a leucocratic dyke is 172.6 ± 1.7 Ma (Dimo-Lahitte *et al.* 2001). A plagiogranite indicates a date of 163.8 ± 1.8 Ma (early Callovian), and doleritic dykes yielded a date of 172.6 ± 1.7 Ma (Aalenian; Vergely *et al.* 1998). At the base of both ophiolitic belts, widespread metamorphic soles dated between 174.0 and 160.0 Ma also occur (Aalenian–Oxfordian; Vergely *et al.* 1998; Dimo-Lahitte *et al.* 2001).

In both ophiolite belts, the radiolarites linked to the volcanics yielded radiolarian assemblages of Middle Jurassic age (Bathonian–Callovian/Oxfordian; Chiari *et al.* 2004b).

Pindos (Greece)

The Pindos ophiolitic nappe represents the southern extension of the Mirdita ophiolitic nappe. Within this nappe, the Avdella mélange constitutes the base of two ophiolite sequences: the Dramala and the underlying Aspropotamos units (Jones & Robertson 1991) of Bathonian–Callovian age (Jones *et al.* 1992). Metamorphic soles occur at the bottom of both sequences.

The Avdella mélange is comparable in lithology and tectonic position to the Rubik complex of the Mirdita ophiolitic nappe. The blocks of basalt included in the mélange indicate WPB, WPB–MORB transitional and MORB compositions (Jones & Robertson 1991). Jones *et al.* (1992) reported the occurrence of radiolarian cherts of Triassic and Jurassic age in isolated blocks and in volcanoclastic successions. Probably, the basalts are both Triassic and Jurassic, as also observed in the mélanges of the Mirdita and Argolis areas.

The Aspropotamos unit includes serpentinites, an intrusive sequence of MOR affinity, and volcanic rocks, which at the base have a MOR affinity; upwards, at the top, boninites are present (Jones & Robertson 1991; Saccani & Photiades 2004).

The Dramala unit shows serpentinized harzburgites and ultramafic cumulates; the latter are cut by boninitic dykes.

The metamorphic soles gave Ar–Ar dates at 165.0 ± 3.0 and 169.0 ± 5.0 Ma (Bajocian–Bathonian; Spray *et al.* 1984) and a K–Ar date at 176.0 ± 5.0 Ma (Toarcian–Aalenian; Thuizat *et al.* 1981; corrected by Spray *et al.* 1984). At the base of the soles, Carnian–Norian ages were found in chert packets, tectonically at the top of slightly metamorphosed basalts, interpreted by Jones *et al.* (1992) as part of the metamorphic sole.

Vourinos (Greece)

The Vourinos ophiolite is the eastern continuation of Pindos and can be correlated to the Dramala unit and the underlying Advella mélange.

The Ayos Nicolaos mélange lies at the base of the ophiolitic nappe, which consists of a very thick sequence of SSZ setting (Beccaluva *et al.* 1984) composed of harzburgites, a gabbroic complex (from dunites to diorites), a sheeted dyke complex, and volcanics (basalts and diorites) that are crosscut by boninitic dykes.

At the top of the sequence, a thin chert level yielded radiolarian assemblages of Bathonian–Callovian age (Chiari *et al.* 2003).

At the base of the sequence, slices of metamorphic soles yielded Ar–Ar dates at 171.0 ± 4.0 Ma (Aalenian–Bajocian boundary; Spray *et al.* 1984).

Koziakas (Greece)

In the Koziakas area, the southeastern continuation of the Pindos zone, two ophiolitic tectonic units crop out. The lower unit, the Koziakas mélange, includes blocks of red cherts – volcanic rocks with occasional chert layers. The volcanic blocks show different magmatic origins: MORB (Capedri *et al*. 1985), transitional to alkaline rocks and boninites (Saccani *et al*. 2003). Radiolarian assemblages found in the cherts linked to the basalts yielded two different ranges of ages: Middle–Late Triassic and Middle–Late Jurassic (M. Chiari, pers. comm., 2004). Since the Koziakas mélange is comparable in lithology and tectonic position to the Rubik complex of the Mirdita ophiolitic nappe of Albania, it can be surmised that the Triassic radiolarian cherts could be associated with the MOR basalts, and the Jurassic cherts are associated with both MOR and transitional basalts and boninites.

The overlying ophiolite unit consists mainly of serpentinized harzburgites and gabbros (Capedri *et al*. 1985), and includes basalt and plagiogranite dykes cut by boninite dykes, as observed in the Vourinos Massif (Saccani *et al*. 2003).

Othris (Greece)

In the Othris zone, immediately to the south of the Koziakas area, three ophiolitic tectonic units crop out. From bottom to top, these are the Agoriani mélange, the Middle harzburgitic unit and the Upper lherzolitic unit. The mélange contains fragments of peridotites, gabbros and basalts somewhere linked to chert layers. The basalts are N-MORB to E- or T-MORB (Lefèvre *et al*. 1993), MORB, ocean island basalts (OIB), MORB-IAT and boninites (Photiades *et al*. 2003), implying that their origin occurred in different oceanic settings. The chert layers yielded late Triassic and Middle Jurassic ages (M. Chari, pers. comm., 2002). No clear stratigraphic relationship was observed between the volcanics and the cherts. However, as observed in Mirdita and Argolis, a Late Triassic age for some of the MOR basalts is suggested.

The metamorphic sole of this ophiolite yielded an Ar–Ar date of 169.0 ± 4.0 Ma (Bajocian; Spray *et al.* 1984); a rhyolite a K–Ar date of 181.0 ± 70 Ma (Sinemurian–Plinsbachian); and a dolerite a K–Ar date of 156.0 ± 6.0 (Oxfordian–Kimmeridgian; Hynes *et al.* 1972).

Evvoia Is. (Greece)

In Evvoia Is., the Pagondas mélange is tectonically covered by extensive ophiolite masses and contains blocks of ophiolites, including basalts (transitional MORB–WPB, Danelian & Robertson 2001) and cherts. The chert blocks yielded two different groups of ages: Middle–Late Triassic and Middle Jurassic. Here, as also in Othris, no physical continuity exists between the studied volcanite and chert sections (see Bortolotti *et al*. 2003). However, basalts of both Triassic and Jurassic age are probably present (Danelian & Robertson 2001).

A radiometric age on the metamorphic sole offered a not completely reliable Ar–Ar date of approximately 180.0 Ma (?Toarcian; K–Ar; Spray *et al.* 1984).

Kerassies–Milia (Greece)

This minor ophiolite belt consist of serpentinites, gabbros and pillow lavas, tectonically interfingered in the Pindos flysch. The basalts have OIB and MORB affinity, and 'some IAT characteristics' (Pe-Piper & Hatzipanagiotou 1993). A radiometric date on gabbros yielded 210.0 Ma (Norian?; Sm–Nd; Pe-Piper 1998).

Argolis Peninsula (Greece)

In the Argolis Peninsula three ophiolitic units crop out: the Migdalitza Ophiolite, covered by a Lower Cretaceous–Eocene 'Mesautochthon'; the Iliokastron mélange (Middle–Late Jurassic–Early Cretaceous) and the Adheres mélange (Cretaceous–Paleocene). In particular, the Migdalitza ophiolite unit consists of MOR basalts (Clift & Dixon 1998; Saccani *et al.* 2004a), either interbedded with, or covered by layered radiolarian cherts. This complex shows scattered serpentinite slivers at the base. The cherts yielded radiolarian assemblages of the Middle and Late Triassic, but also of the Early and Middle Jurassic (Bortolotti *et al.* 2003, 2004c).

The Iliokastron mélange includes blocks of serpentinized harzburgites, basalts and boninitic rocks. In the Adheres mélange, radiolarian cherts at the top of basalts yielded a radiolarian assemblage of the Middle–Late Jurassic (Bortolotti *et al*. 2003).

Southern Peloponnese (Greece)

Near Angelona, a small ophiolitic complex crops out, consisting of peridotites (with a metamorphic sole at the base) and gabbros covered by a chert level. This level yielded a radiolarian assemblage of Carnian–Norian age (Danelian *et al*. 2000).

Guevgueli–Chalkidiki (Greece)

The Guevgueli–Chalkidiki ophiolitic complex, the easternmost ophiolite of Greece, consists of discontinuous bodies of harzburgites and dunites, gabbroic rocks, sheeted dykes and basaltic lavas, of an SSZ (low-Ti) environment (Christodoulou & Michaelides 1990). A hornblende gabbro of the Thessaloniki ophiolite yielded a radiometric date of 172.0 ± 5.0 Ma (Aalenian–Bajocian; Kreuzer in Mussallam & Jung 1986). K–Ar datings of a gabbro gave 149.0 ± 3.0 Ma (Kimmeridgian–Tithonian boundary); of a diorite 163.0 ± 3.0 and 154.0 ± 3.0 Ma (Callovian–Kimmeridgian; Spray *et al.* 1984). Radiolarian assemblages of Oxfordian age have been found in siliceous sediments at the top of the basalts by Danelian *et al.* (1996).

At Guevgueli the ophiolite is cut by the potassium-rich Fanos granite, dated by Borsi *et al.* (1996) at 150.0 ± 2.0 Ma (Tithonian–Kimmeridgian boundary) with eight Rb–Sr (147.0, 148.0, 149.0, 151.0, 151.0, 152.0, 153.0, 153.0 Ma) and one K–Ar (150.0 Ma) dates. Spray *et al.* (1984) found 148.0 ± 3.0 Ma (Tithonian; K–Ar). Southwards, the similar Monopigadhon granite gave a date of 149.0 Ma (Kreuzer in Mussallam & Jung 1986).

Crete Is.

The ophiolites of Crete consist of a composite ophiolitic nappe, characterized by an ophiolitic mélange at the base (including both basalts and metamorphic soles), and an ophiolite complex (mainly serpentinized lherzolites cut by gabbroic dykes at the top) (Seidel *et al.* 1981). In the mélange, isolated basalt bodies show MORB affinity (Koepke *et al.* 2002; references therein).

Seidel *et al.* (1981) report K–Ar dates of gabbroic rocks of the ophiolite nappe between 135.3 ± 3.5 Ma and 156.2 ± 3.1 Ma (Callovian–Valanginian), and of the metamorphic rocks of the mélange, some of which are ophiolitic, between 78.8 ± 8.0 Ma and 67.6 ± 1.8 Ma (Campanian–Maastrichtian).

Further K–Ar dates are reported by Koepke *et al.* (2002): (i) a body of hornblendites associated with the serpentinites (metamorphic sole?) from 166.0 ± 3.0 to 148.6 ± 2.1 Ma (Bathonian–Kimmeridgian); (ii) the date of gabbroic dykes varies from 120.0 to 160.0 Ma, with a cluster at approximately 140.0 Ma (Berriasian–Valanginian); and (iii) the metagabbros are approximately 95.0 Ma (Cenomanian). The authors conclude that the ophiolites that pertain to the Balkan Belt (the 160.0 Ma date) were intruded by gabbroic dykes 20 my later (after the 140.0 Ma cluster) and underwent orogenic metamorphism at approximately 95.0 Ma .

Chios Is. (Greece)

In the island of Chios, a minor ultramafic ophiolite mass of SSZ environment with metamorphic soles at the base crops out (Pickett & Robertson 1996). In western Chios, at the border of the Sakarya microcontinent, high-Ti subalkaline basalts gave a Sm–Nd date of 210.0 Ma (Pe-Piper 1998).

Lesvos Is. (Greece)

In Lesvos Is., an ophiolitic mélange, includes serpentinites, gabbros and basalts (Pe-Piper *et al.* 2001; Hatzipanagiotou *et al.* 2003; bibliography therein). The basalts have MORB and transitional MORB-IAT affinity. The ophiolite nappe on top of the mélange consists of serpentinized harzburgites and lherzolites, with a metamorphic sole, which shows WPB and N-MORB affinity (Pe-Piper *et al.* 2001). The subophiolitic metamorphic sole yielded K–Ar dates of 158.0 ± 5.0 and 153.0 ± 5.0 Ma (Oxfordian; Hatzipanagiotou & Pe-Piper 1995).

Northern and central Turkey

In northern and central Turkey, several ophiolite complexes and a widespread ophiolitic mélange occur. Unfortunately, age data are lacking for most of them; therefore, the few available data will be reported here.

In the Sakarya area, the central Sakarya ophiolitic complex can be subdivided in two units: the Daðküplü mélange at the base, and the Taþkepe Ophiolite at the top (Tekin *et al.* 2002). Blocks of basalt, the dominant lithology of the mélange, range in composition from IAT to MORB, OIB and calc-alkaline basalts, and were clearly formed in different oceanic settings (and ages?). In a large block, thin layers of radiolarian cherts alternated with basalts that yielded radiolarian assemblages of late Carnian age (Tekin *et al.* 2002). Middle–Late Jurassic ages were also found in other chert outcrops (Göncüoðlu *et al.* 2000).

The Ankara mélange includes blocks of ophiolitic rocks (serpentinites, gabbros and volcanics, linked sometimes to cherts with radiolarians). The radiolarian cherts studied by Tekin (1999) yielded a radiolarian assemblage of late Norian age. Other isolated chert blocks, studied by Bragin and Tekin (1996), yielded Upper Triassic radiolarian assemblages. In the same area, Lower Jurassic (Kimmeridgian–Tithonian) and Lower Cretaceous cherty blocks were found by the same authors.

Near Kure, in the Central Pontides, a dismembered ophiolite composed of slices of serpentinites, gabbros, a sheeted complex and basalts (MOR and IAT) pertains to the Kure complex (Ustaömer & Robertson 1997). K–Ar dating of basalts indicates a Middle Jurassic age (Aydin *et al.* 1995 in Ustaömer & Robertson 1997). However, according to Ustaömer & Robertson (1997), the data are not reliable, only because they consider the Kure complex a Triassic to Lower–Middle Jurassic subduction–accretion complex.

Summarizing the previous observations, it can be concluded that (i) The MORB ophiolites show ages between Middle Triassic (?Anisian) and Kimmeridgian, although Early Jurassic ages are very rare; (ii) the metamorphic soles formed between the ?Toarcian and the Kimmeridgian, deriving from older oceanic crust; (iii) the deep thrusting that produced the metamorphic soles was intra-oceanic, and it occurred during the formation of the basalts and cherts; (iv) the IAT rocks are only Middle–Late Jurassic in age; and (v) The closure of the Western Tethys began in the Middle Jurassic and was completed around the Jurassic–Cretaceous boundary.

Cretaceous ophiolites

Scattered from the southeasternmost Greek Is, through northwest Turkey, the Taurus Belt (southern Turkey) and Cyprus, to northern Syria, a series of Cretaceaus supra-subducion ophiolitic units of very variable dimensions crop out. They are considered to be the remnants of a post-Pangea break-up Cretaceous Neotethys sited south of the Menderes Massif. Accordingly, they will not be considered in the present paper. For an updated synthesis of the studies and knowledge, see Dilek *et al.* (1999) and Robertson (2002).

Figure 9 quotes the chronological data found in the literature.

Summarizing these data: (i) except for some scattered and dubious cases, the ophiolites were

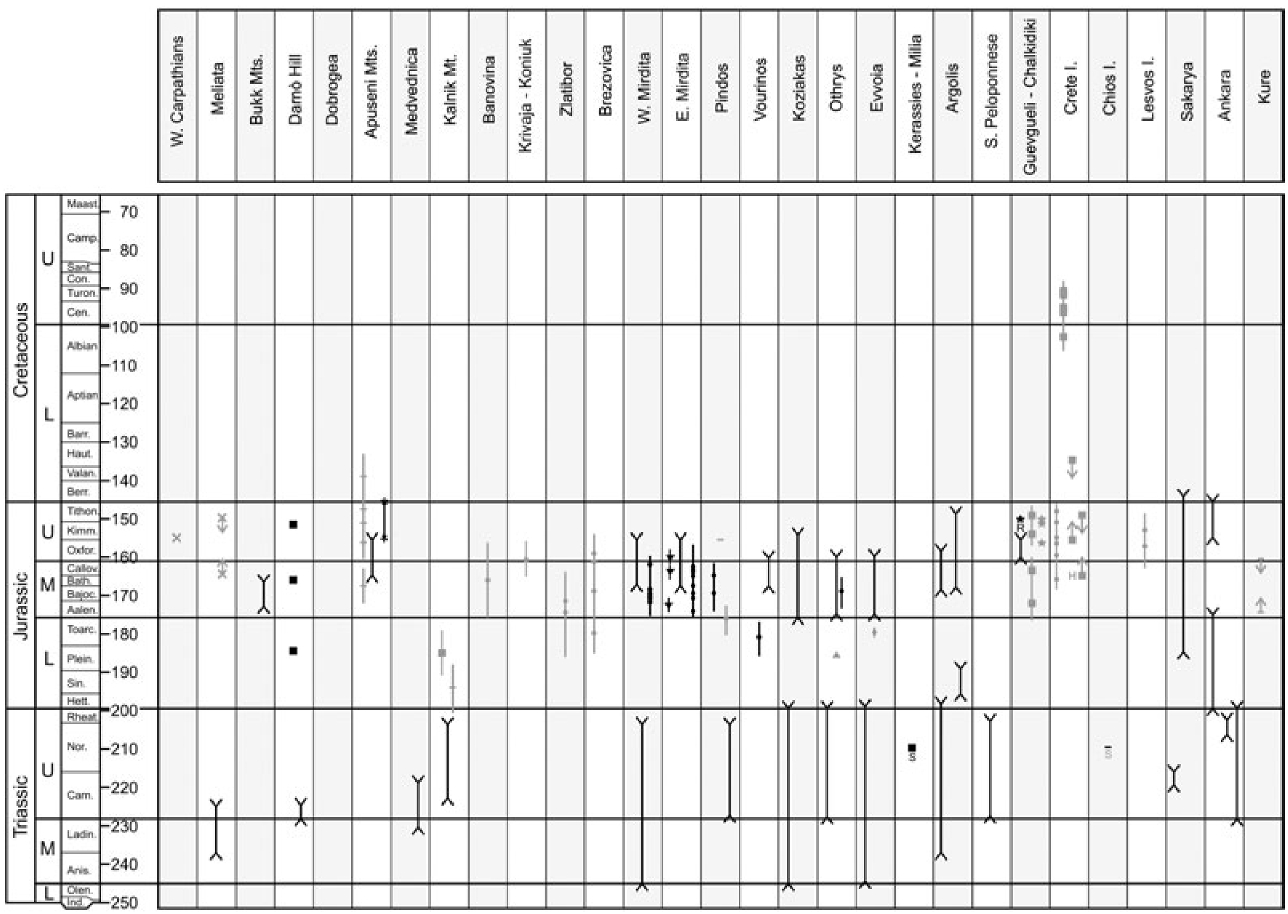

Fig. 9 Chronological data on the Triassic–Jurassic Eastern Tethys ophiolites. L, Lower; M, Middle; U, Upper. Radiometric ages: black, Ar–Ar; gray, K–Ar; S, Sm–Nd; U, U–Pb; (×), blueschist ophiolites; (■), gabbroic rocks (H, hornblendite); (-), basaltic rocks; (●), metamorphic soles; (▼), dykes; (▲), rhyolites; (★), granitic rocks. Biostratigraphic ages: (*), foraminifers; (I), radiolarians.

formed during the uppermost Early Cretaceous and the Late Cretaceous; (ii) the metamorphic soles, cut locally by gabbro dykes, are roughly of the same age; (iii) the deep thrusting that produced the metamorphic soles was here, as in the Eastern Tethys, intraoceanic; and (iv) the presubduction MORB crust has been completely subducted – it was older than Late Cretaceous, probably Early Cretaceous, in age. Thus, one might be the age of the birth of the Mediterranean Neo-Tethys.

WESTERN TETHYS

In the Western Tethys domain (Fig. 3), only Middle–Upper Jurassic ophiolites carrying a MORB signature are known. From east to west, MORB ophiolites are found in the Alps, in Corsica, in the Northern and Southern Apennines, in scattered outcrops in northwestern Africa and in the Betic Cordillera of southern Spain (Principi *et al.* 2004; references therein). Previous interpretations suggested that these ophiolites are derived from a single oceanic branch of the Tethys: the Ligurian–Piedmontese Ocean, or Western Tethys.

In the Alps, most ophiolite rocks suffered blueschist and eclogitic orogenic metamorphism, as did those cropping out in the Betic Cordillera. A portion of the ophiolites of Corsica and the Southern Apennines was also subjected to the same type and degree of metamorphism.

Below, the main and better known ophiolitic areas are reviewed. Figure 9 quotes all the chronological data found in the literature.

Alps

The easternmost 'Alpine' ophiolite crops out in the Rechnitz window, east of Vienna; to the west, the metamorphic Alpine sequences border the nonmetamorphic sequences of the Northern Apennines along the Sestri–Voltaggio Line (see Abbate *et al.* 1970).

1. In the Central Alps, ophiolites that are considered to have been formed next to the continental margin (transition ocean–continent as in the Galician margin, see Desmurs *et al.* 2001) crop out. They consist of peridotites, gabbro bodies and basalt flows, with scattered layers of cherts and pelagic carbonates. The U–Pb date of a gabbroic body is 161.0 ± 1.0 Ma (Callovian–Oxfordian boundary; Schaltegger *et al.* 2002). Peters and Stettler (1987) found an age of 160.0 ± 8.0 Ma (Callovian–Oxfordian boundary) for the cooling of an ultramafic body.
2. In the Western Alps, several ophiolite bodies with either complete or partially complete oceanic successions occur. Fewer masses of percolated subcontinental lherzolites, e.g. at Lanzo, Piccardo *et al.* 2004) also occur, but their ages are unknown. Geochemical data point to a subcontinental origin of the basal mantle rocks, whereas the overlying magmatic portion (gabbros and basalts) has a MORB signature.

Four chert sections on top of the basalts yielded radiolarian assemblages indicating the following ages: middle–late Oxfordian at Saint Veran and late Bathonian–early Callovian at Traversiera (De Wever & Baumgartner 1995); late Oxfordian–early Kimmeridgian at Chabrière (Schaaff *et al.* 1985); and middle Bathonian at Col de Gets (in a chert block associated with the ophiolitic rocks; Bill *et al.* 2001). In general, these ages indicate that radiolarian cherts formed between the middle Bathonian and late Kimmeridgian.

Only a few radiometric dates are available: metagabbros from Antrona ophiolite gave sensitive high mass-resolution ion microprobe zircon dates of 155.6 ± 2.1 and 155.0 ± 2.0 Ma (Oxfordian–Kimmeridgian boundary; Liati *et al.* 2003); gabbro blocks from the Gets nappe yielded 166.0 ± 1.0 Ma (U–Pb) and 165.9 ± 2.2 Ma (Ar–Ar; early Bathonian; Bill *et al.* 1997); an eclogitic metagabbro from Zermatt-Saas yielded a date of 163.0 ± 2.0 Ma (Callovian; U–Pb; Rubatto & Gebauer 1996). A similar metagabbro in the Ligurian Alps yielded a Sm–Nd date of 177.0 ± 23.0 Ma (Borsi 1995). In the Montgenèvre ophiolite, a dioritic vein within a gabbro yielded a date of 156.0 ± 3.0 Ma (Oxfordian) and an albitite within the mantle rocks yielded a date of 148.0 ± 2.0 Ma (Tithonian; U–Pb; Costa & Caby 2001).

Two plagiogranites and the host Fe-diorite from the Ligurian Alps yielded radiometric dates from 150.0 ± 1.0 to *ca* 156.0 Ma (Kimmeridgian; U–Pb on zircon, Borsi *et al.* 1996).

Corsica

In Corsica, the ophiolites comprise mantle lherzolites, gabbros and basalts with (N- and T-) MORB signatures (Saccani 2003; bibliography therein). Cherts overlying the basalts yielded radiolarian assemblages of late Bajocian to early Callovian (De Wever & Danelian 1995; Chiari *et al.* 2000). A U–Pb date of 161.0 ± 3.0 Ma for a plagiogranite body (Callovian–Oxfordian boundary) has been reported by Ohnenstetter *et al.* (1981).

Northern Apennines

The ophiolites of the Northern Apennines of Italy occur in two different tectonic positions: (i) in the basal section of the Vara Supergroup; and (ii) in olistoliths in Cretaceous–Eocene preflysch and flysch units. Most of the ophiolite successions are incomplete and composed of serpentinized lherzolites, scarce gabbros, and basalts, with MORB signatures (Beccaluva *et al*. 1989), in thin, very discontinuous layers. Ophiolitic breccias linked to the basalts are wide spread (Abbate *et al*. 1980). The radiolarian assemblages of the cherts at the top of the ophiolites cluster into two age groups: (i) in the northern portion of the chain (Eastern Liguria), a thin layer of cherts at the base of the basalts was found to be Bajocian in age and at the top of the basalts the age was late Bathonian; (ii) in the southern portion (Tuscany), a thin layer of radiolarian cherts at the base of a section of basalts indicated a Bathonian–Callovian age. All the chert sections at the top of the basalts gave ages ranging between Callovian and Kimmeridgian (Chiari *et al*. 2000).

The Sm–Nd isochrons determined on plagioclase–clinopyroxene pairs from a peridotite of the Ligurian–Emilian Apennines indicate a date of 164.0 ± 20.0 Ma (Middle–Late Jurassic), which has been interpreted as the time of the plagioclase facies re-equilibration (Rampone *et al*. 1995).

In Eastern Liguria (Val Graveglia), a gabbro gave a Sm–Nd date of 164.0 ± 14.0 Ma (Aalenian–Oxfordian; Rampone & Hofmann 1998). In the same area, a plagiogranite in ophiolitic breccia underneath the basalts yielded a date of 153.0 ± 0.8 Ma (Kimmeridgian; U–Pb zircon; Borsi *et al*. 1996). In the Sestri–Voltaggio zone, at the boundary with the Ligurian Alps, a plagiogranite yielded a date of 153.3 ± 1.0 (Kimmeridgian; U–Pb; Borsi *et al.* 1996). Southwards, in Tuscany, basalts and plagiogranites intruded into the basalts yielded dates of 157.0–158.0 Ma (Oxfordian; Ar–Ar; Bortolotti *et al.* 1995).

Southern Apennines

In the Southern Apennines, metamorphic and nonmetamorphic ophiolitic successions having MORB signature crop out (Lanzafame *et al*. 1979; Beccaluva *et al*. 1982; Spadea 1994).

A non-metamorphic section yielded a radiolarian assemblage of middle Callovian–early Oxfordian age in a thin chert layer at the top of the basalts (Marcucci *et al*. 1987).

Sicily–Rif

In northeastern Sicily, some olistoliths of basic rocks occur in the chaotic 'Argille Scagliose'. Their age is unknown. Basic volcanic rocks, associated with radiolarites and/or limestones of Late Dogger–Early Malm age, are found as tectonic slices or olistoliths in the Maghrebian flysch formations of north Africa.

In both areas, the basic rocks show E-MORB characteristics: they would have formed in small oceanic scars, scattered along the wrench zone, which linked the Western Tethys to the Central Atlantic (Durand-Delga *et al.* 2000; references therein).

Betic Cordillera

In the Betic Cordillera, a tectonic unit of high-pressure metamorphic ophiolites consists of serpentinized harzburgites, MORB troctolites, gabbros and basalts transformed to eclogites and amphibolites (Puga 1990). The sedimentary cover is mainly calc-schists and quartzites. The only available age data are 213.0 ± 2.5 Ma (Norian) for a metagabbro and 158.0 ± 4.5 (Oxfordian) for a brown amphibole vein in a metabasalt (Ar–Ar; Puga *et al.* 1993, 1995).

Summarizing the previous data: (i) the magmatic section of the ophiolites has a MORB signature and the mantle rocks are mainly of subcontinental origin; (ii) no metamorphic soles or suprasubduction rocks are present; and (iii) the opening of the ocean seem to be slightly older in the northern portion (Bajocian) than in the southern portion (Oxfordian).

CENTRAL ATLANTIC OCEANIC CRUST

The Atlantic Ocean can be subdivided into three sections, which were formed at separate times. From oldest to the youngest they are: the Central, South and North Atlantic. The ideal boundary between the North and South Atlantic can be located along the Azores–Gibraltar fracture zone, which, at its western side, corresponds to the Newfoundland fracture zone (NFZ; see Srivastava *et al*. 1990). The boundary between the Central and South Atlantic is marked by the Romanche fracture zone.

The focus in the present paper is on the older oceanic crust of the Central Atlantic, whose initial spreading is contemporaneous with the opening of the Western Tethys.

The oldest ocean floor occurs along the western border of the Central Atlantic, near the Blake–Bahama Basin. Immediately to the east of the Blake Spur anomaly, Bryan *et al.* (1980) hypothesized the existence of anomaly M28 near Deep Sea Drilling Project site 391. At Ocean Drilling Project site 534, located 15 nautical miles to the southwest, radiolarian oozes at the top of the basalts yielded radiolarian assemblages dated as middle Bathonian (Baumgartner & Matsuoka 1995). This indicates that the Central Atlantic opened at approximately 170.0 Ma.

THE PERI-CARIBBEAN DOMAIN

In the Caribbean region, Jurassic and Cretaceous MORB ophiolites are described in many localities (Fig. 10).

Hispaniola

In the Duarte metaophiolitic unit (Draper & Lewis 1989), radiolarian assemblages in chert levels associated with ophiolites provided an Early Oxfordian–Tithonian (Aguacate) and a Jurassic–Cretaceous age (Janico) (Pessagno 1993 in Montgomery & Pessagno 1994).

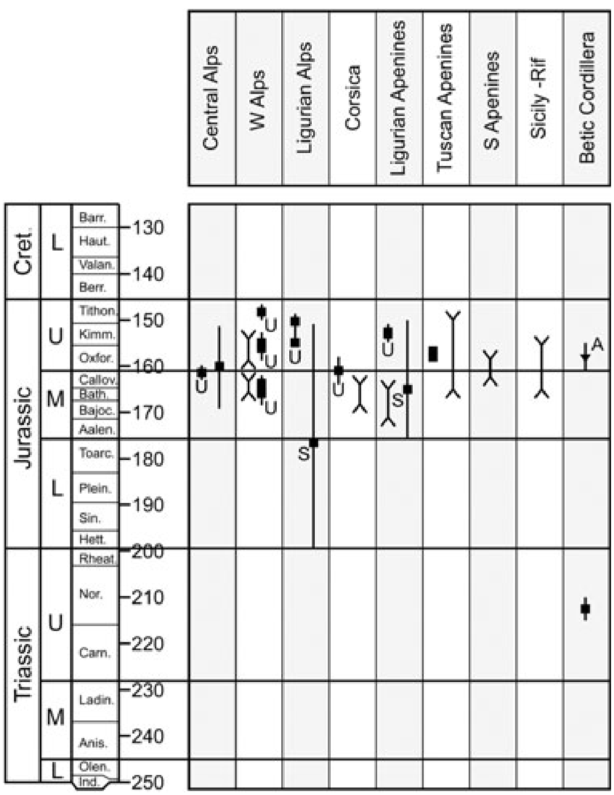

Fig. 10 Chronological data on the Jurassic Western Tethys ophiolites. A, amphibole vein in basalts; Cret, Cretaceous; L, Lower; M, Middle; U, Upper. For other symbols, see Figure 9.

Puerto Rico

According to Montgomery *et al.* (1994a, 1994b) in the ophiolitic mélange of the Bermeja complex, radiolarian cherts linked to the ophiolites gave Pliensbachian, Callovian–Oxfordian, Oxfordian–Tithonian, Hauterivian–Barremian, Aptian, Cenomanian and Turonian ages.

Desirade Is.

Tithonian cherts and limestones interbedded with pillow basalts are documented in Desirade Is. (Montgomery *et al.* 1992). Coeval MORB plagiogranites (145.0 ± 5.0 Ma; Tithonian–Berriasian; Mattinson *et al.* 1980) are also present.

Cuba

Ophiolites occur in the Northern Cuba Fold Belt, in the Allochthonous terranes and in the Northern Ophiolitic mélange (Kerr *et al.* 1999; references therein).

Northern Cuban Fold Belt

Within the Placetas Belt, the cherts associated with the pillow basalts have a Tithonian–Maastrichtian age (Iturralde-Vinent & Mari Morales 1988). According to Iturralde-Vinent and Mari Morales 1988, these rocks formed during the rifting phase.

Allochthonous terranes

The El Sàbalo Formation consists of a thick sequence of pillow basalts, hyaloclastites and limestones of late Oxfordian–early Kimmeridgian age (Kerr *et al.* 1999). Iturralde-Vinent (1988, 1996a) considered these basalts to have formed in a continental rift.

In the cherts of the Encrucijada Formation – a fragment of the ocean crust of the Caribbean within a back-arc–marginal sea environment – Cruz and Simón (1992, in Kerr *et al.* 1999), Iturralde-Vinent (1994), Aiello and Chiari (1995), Iturralde-Vinent (1996b), Cruz (1998) and Aiello *et al.* (2004) found a middle Albian–Cenomanian radiolarian assemblage.

Northern ophiolite mélange

In the Northern ophiolite mélange, ophiolitic blocks belonging both to a ?Jurassic–?Cretaceous oceanic domain and to the Cretaceous island arc unit are present (Iturralde-Vinent 1994). The mafic and sedimentary sections correlated to those of the Encrucijada Formation and, likewise, have been interpreted as having been formed within a marginal sea–back-arc (Iturralde-Vinent 1994, 1996c), or within a supra-subduction forearc environment (Andó *et al.* 1996).

In Central Cuba, at Mina Margot, the volcanic-sedimentary section comprises pillow basalts covered by sediments (Margot Formation) of Aptian–Albian age (Kerr *et al.* 1999).

In the same area, the Sagua la Chica Formation consists of basalts intercalated with sediments and tuffaceous rocks. The radiolarian assemblages found in the formations indicate a Tithonian age (Llanes *et al.* 1998). Iturralde *et al.* (1996) provided ambiguous K–Ar dates for a dolerite cropping out near Holguin, of 126.0 ± 8.3, 102.0 ± 20.0, 98.2 ± 5.0 and 57.8 ± 5.4 Ma (Albian–Paleocene). Kerr *et al.* (1999) classified these rocks as boninites.

In Eastern Cuba, the volcanic-sedimentary sections have been dated as Tithonian–Campanian (Iturralde-Vinent 1996a).

Kerr *et al.* (1999), discussing the more recent studies on the Cuban ophiolites, conclude that only the ophiolites in the Encrucjiada Formation represent remnants of the proto-Caribbean oceanic basin.

Guatemala

In the Motagua suture zone, the meta-ophiolitic sequence of the Tambor Group (Beccaluva *et al.* 1995; Giunta *et al.* 2002; references therein) consist of blueschist metabasalts, of MORB affinity, covered by metacherts and micaschists (Giunta *et al.* 2002; bibliography therein). In the southern sector of the suture zone, along the Rio Grande, south of Puente del Rio Grande, deformed metacherts yielded radiolarian associations of Late Jurassic age (Chiari *et al.* 2004a,b).

Costa Rica

Ophiolitic Nicoya complex

Galli Oliver (1977, in Baumgartner 1987) and Schmidt-Effing (1979, in Baumgartner 1987) referred the cherts of the complex to the Tithonian–Valanginian age.

Kuijpers (1980) subdivided the complex in two main units: the Matapalo and the overlying Esperanza unit. The radiolarian assemblages yielded Berriasian to Aptian ages in the Matapalo unit, and Cenomanian to Early Santonian in the Esperanza unit. Cenomanian radiolarian assemblages were found in the Matapalo unit by Schmidt-Effing (1980).

Jurassic (Pliensbachian?, Callovian–Oxfordian) and Cretaceous (Neocomian–Barremian, Valanginian–early Albanian, early Albian, Barremian–Cenomanian) ages were found by De Wever *et al.* (1985) in the Volcano-Sedimentary Formation. Callovian–early Oxfordian radiolarian assemblages were found by Baumgartner (1984, 1995) in the cherts at the top of the basalts, near Santa Rosa.

The ages of the cherts associated with the basalts of the Nicoya complex seem to range from ?Pliensbachian to Santonian.

Herradura block

Baumgartner *et al.* (2000) and Popova *et al.* (2000) dated the cherts of the sedimentary cover of the Herradura block (gabbros and basalts, central Costa Rica) to the Campanian (?early Maastrichtian).

Venezuela

At Sisquisique-Loma de Hierro, Venezuelan coastal range, MORB pillow lavas are linked in a matrix with Bajocian ammonites (Stephan 1980, in Beck *et al.* 1984; Beck *et al.* 1984). In contrast, Girard *et al.* (1982) considered the basalts to be of early Late Cretaceous age, and of a plateau magmatism.

To summarize the preceding data: (i) magmatic MORB sections of Late Jurassic age are present; and (ii) the older ages of the radiolarian cherts seem to be Plienslachian, but Late Jurassic–Early Cretaceous ages are more common. This might be the age of the birth of the Caribbean Tethys.

DISCUSSION

RIFTING

It is commonly accepted that the beginning of the rifting stage that affected the future continental margins of the Mesozoic ‘ophiolitic basins’ in the peri-Mediterranean area (Fig. 4) and in the future Central Atlantic Ocean (Fig. 6) is Anisian–Ladinian.

The rifting stages in the peri-Caribbean zone also developed during the Triassic and Jurassic, even though this is less documented. In particular, Triassic rifting structures are documented north of the Gulf of Mexico. According to Bartok (1993), the rifting to the south developed during Jurassic times. Triassic rifting is probably linked to the opening of an oceanic basin in the Gulf of Mexico – the Jurassic rifting to the proto-Caribbean ocean (Emery & Uchupi 1984; Ross & Scotese 1988; Case *et al*. 1990; Stephan *et al*. 1990).

It is also worth noting that, during the Triassic, magmatism accompanied rifting in other areas of Pangea: in the area of the future North Atlantic (Malod & Mauffet 1990), between Africa and Arabia to the north, and between Madagascar and India to the south (see Fig. 2). In conclusion, it emerges that the framework of continental rift basins, along which sea floor spreading and the break-up of Pangea occurred, developed mainly during the Triassic.

SEA FLOOR SPREADING

The formation of the peri-Mediterranean (intra-Pangean) Tethyan Ocean began in the Middle–Late Triassic in the eastern peri-Mediterranean region (Eastern Tethys); in the Middle Jurassic in the western peri-Mediterranean region (Western Tethys) and the Central Atlantic, and in the Late Jurassic or Early Cretaceous in the peri-Caribbean area. Here, the peri-Mediterranean area, and the area westwards to the Caribbean, will be examined, and some of the essential data for the Central Atlantic Ocean will be pointed out.

Peri-Mediterranean region

The ophiolites of the eastern and western parts of the peri-Mediterranean region show significant differences concerning both their petrology and their age. In this area, three different types of ophiolitic successions can be recognized: (i) MORB and IAT, Triassic and Jurassic in eastern areas (the Eastern Tethys); (ii) IAT, Late Cretaceous in southeastern areas (the Neotethys); and (iii) MORB, Jurassic in western areas (the Western Tethys).

Eastern Tethys

The long and complex oceanic evolution of the Eastern Tethys is well documented. A Middle–Upper Triassic phase of sea floor spreading is documented by the age of the radiolarian cherts above MOR basalts. These Triassic oceanic remnants occur all along the ophiolitic belt of the peri-Mediterranean region: from the Western Carpathians to the north, throughout the Dinarides and Hellenides, to the Pontic Ranges to the southeast. The age of radiolarian cherts in the top MOR basalts indicates that drifting continued until the Middle to earliest Late Jurassic. During the Middle Jurassic, the onset of IAT suggests that the subduction of the oceanic crust had already started by the Middle Jurassic, probably during uppermost Early Jurassic–earliest Middle Jurassic time. During the Late Jurassic, the oceanic areas seem to have already closed, as indicated by Middle Jurassic blocks of continental origin included in the mélanges underlying the ophiolites and by the Tithonian age of the mélanges and flysch that unconformably cover the ophiolitic sequence.

How many oceans in the Eastern Tethys domain?

In the past few years, different reconstructions of the Mesozoic evolution of the Dinaric–Hellenic ophiolite successions have been proposed. These hypotheses have produced different models for the Triassic–Jurassic paleogeography and geodynamics of the Western Carpathians–Hellenides–Dinarides internal domains. One group of scholars supports the idea that the ophiolites developed in two or more ocean basins, formed contemporaneously or subsequently (e.g. Vergely 1984; Channell & Kozur 1997; Pamić *et al*. 1998, 2002; Ricou *et al*. 1998; Stampfli & Mosar 1999; Robertson & Shallo 2000; Robertson 2002; Stampfli *et al.* 2002). A second group supports the opposite hypothesis that, even if hypothesizing different paleogeographic frames, all Eastern Tethys ophiolites were generated from a single ocean basin (Abbate *et. al.* 1980; Dercourt *et al*. 1993; Dal Piaz *et al*. 1995; Bortolotti *et al*. 1996, 2003, 2004c).

Figure 11 shows that removing the Upper Cretaceous–Tertiary tectonic deformation, all Triassic–Jurassic ophiolites of the Eastern Tethys domain must be shifted oceanwards along the black arrows, forming a single broad alignment that extends between the Western Carpathians and the Pontic Belt, including the Apuseni and the Dinaric–Albanian–Hellenic chains.

All along this alignment, the ophiolites show a similar oceanic evolution during the Triassic–Jurassic. Moreover, the inception of the ocean closure, accompanied by widespread production of IAT magmas, is also almost contemporaneous for these ophiolites (Middle–early Late Jurassic). In the light of this general correlation, the authors of

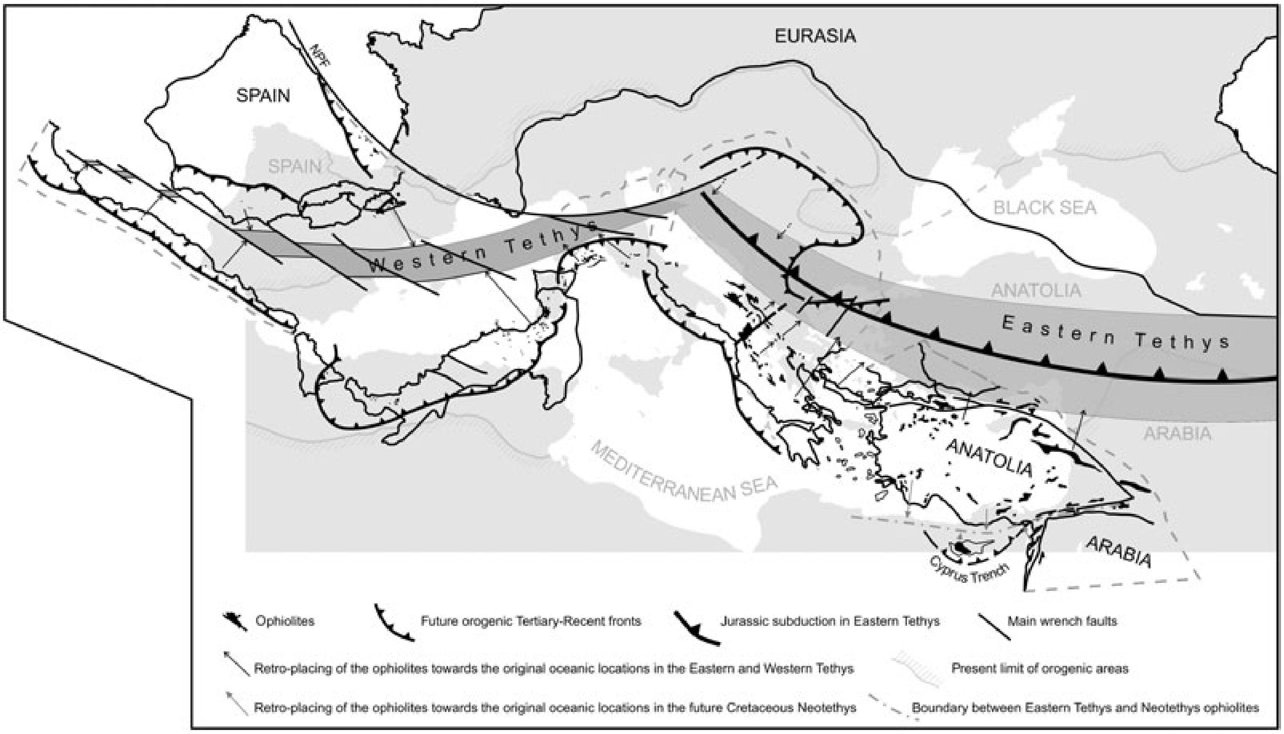

Fig. 11 Palinspastic schematic reconstruction of the peri-Mediterranean area during the Middle–Late Jurassic. Geography: light gray, present-day; black, Middle–Late Jurassic. NPF, North Pirenean Fault.

the present study are inclined to believe that the ophiolitic belt developed in a unique ocean basin. The authors of previous studies, who postulated that the Eastern Tethys ophiolites were derived from different branches of the Tethyan Ocean, based their hypothesis on one main observation: the occurrence, in the Dinaric–Hellenic areas, of two ophiolitic belts – the Vardar zone and the Dinaric zone, up to 180 km apart. While these studies underplay the role of the horizontal translations of the ophiolites, they assume a semi-autochthonous evolution from two or more ocean branches. It is worth noting that where the distance between the two ophiolite belts appears to be the largest (i.e. between the Pindos and the Vardar [or Guevgueli–Chalkidiki] areas; 6–14 of Fig. 3), several tectonic units including ophiolitic masses occur in between: the Vourinos ophiolites, the eastern continuation of the Pindos; the Vermion ophiolites, the Almopias ophiolites and the Vardar ophiolites. All these units occur in a tectonic pile and form westward-dipping thrusts. Moreover, to the north of Greece, in the Mirdita zone, the amphibolite soles show kinematic indicators that suggest an eastern origin for the ophiolitic nappe, which is the eastern side of the Pelagonian units.

Western Tethys

In the Western Tethys domain, sea floor spreading began during the early Middle Jurassic and ended during the Late Jurassic (Tithonian). This is a very short-lived sea floor spreading (between 15 and 20 my), characterized by a very slow rate of drifting (Principi *et al.* 2004; references therein). Ophiolites belonging to this domain are found in the Alps, Apennines, Maghrebid–Rifean and Betic Orogenic Belts. The paleogeography of the Jurassic ocean basin was characterized by an elongated shape oriented WSW–ENE. It is not clear what the location of the eastern end of this oceanic scar was. The evidence indicates that it was probably located near the area presently occupied by the Western Carpathians, and its eastern extension was perhaps touching the western termination of the Eastern Tethys (Fig. 11). The western end of this oceanic basin was probably represented by fewer small and narrow basins that acted as a link between the Western Tethys and the Central Atlantic, along a major wrench structure (the Azores–Gibraltar fracture zone?). The remnants of these small basins are probably represented by the small outcrops of the Rifean and Betic ophiolites.

The Western Tethys remained a fossil basin throughout Late Cretaceous–Early Paleocene time, after which it began to close.

If the Western Tethys ended the spreading phase during the Late Jurassic, and if, by this time, the Eastern Tethys was almost completely subducted, a major question arises: how could the break-up between Laurasia and Gondwana have occurred during latest Jurassic? This question will be addressed in the next paragraphs.

Atlantic Ocean

In the Central Atlantic, the spreading stage began during the Bathonian (Blake Spur anomaly) and is continuing today. During the Late Jurassic, sea floor spreading proceeded from south to north only in the Tagus Plain basin (Malod & Mauffet 1990). The complete separation between Eurasia and North America was completed only during the Tertiary.

Peri-Caribbean area

According to Montgomery and Pessagno (1994), the great abundance of radiolarians *Pantanellidae* in the Jurassic radiolarites of the ophiolite units of the Caribbean region indicate that sedimentation occurred at boreal latitudes. This reconstruction is therefore incompatible with the water temperatures expected in the peri-equatorial Atlantic Ocean, where these radiolarians are rarely found. Montgomery and Pessagno (1994) support the hypothesis that the Caribbean radiolarites, together with the Caribbean oceanic lithosphere, originated to the southwest, in the Pacific Ocean.

Montgomery *et al.* (1994a) dated as Pliensbachian–middle Toarcian the radiolarian cherts above the ophiolitic sequences of Puerto Rico. This is too old, whether these ophiolites were formed in the eastern margin of the peri-Caribbean region or in the Central Atlantic ocean, where the oldest age is Bathonian. Thus, this biostratigraphic result fits more with a Pacific origin for these ophiolites.

According to Montgomery *et al.* (1994a) and Pessagno *et al.* (1999), the peri-Caribbean ophiolites derived from the boreal areas of the Pacific domain, as also suggested by the geodynamic model proposed by Pindel and Barrett (1990). The break-up of Pangea in this region would have been realized mainly throughout strike-slip tectonics, and only a small ocean would have formed in the Gulf of Mexico area (Anderson & Schmidt 1983; Stephan *et al.* 1990).

A different view is proposed by other authors (Case *et al.* 1990; Stephan *et al.* 1990) that consider the peri-Caribbean MORB ophiolites to be remnants of a paleo-Caribbean Tethys.

In the opinion of the authors of the present study, and in disagreement with Pessagno's paleolatitudinal and age interpretation of the Caribbean radiolarites, at least part of the peri-Caribbean ophiolite belt (the ?Jurassic–Cretaceous ophiolites of Cuba and the ?Upper Jurassic ophiolites of the Motagua and the Venezuelan Belts) could have originated from a paleo-Caribbean Tethys.

As also demonstrated by M. Chiari *et al.* (pers. comm., 2003), it is very difficult to hypothesize an extra-Caribbean origin for the Motagua ophiolites, which are now sandwiched between the Chortis and the Maya continental blocks, in a suture zone linked to a probable Jurassic left-lateral transform faulting (Giunta *et al.* 2002). A paleogeographic restoration of the Maya and Chortis blocks should place the latter southwest of the Motagua ocean basin placing this latter in a clearly intrapaleo-caribbean position.

In addition, for the ophiolites of western and Central Cuba, present northwards (at least in part) from the continental metamorphic terranes of Guaniguanico, Pinar del Rio and Camaguey, it is possible to bring analogous considerations, even if with different contexts and tectonic–geodynamic polarities.

Therefore, the oldest ophiolites in the Caribbean region (Late Jurassic and Early Cretaceous) indicate the age of the opening of the Caribbean Tethys.

In this case, it is possible that at least some Jurassic ophiolites of the eastern border of the Lesser and Greater Antilles could pertain to the Central Atlantic oceanic crust, partly obducted and partly subducted.

According to this hypothesis, the break-up was realized through ocean basins developed with a left lateral transtensive peri-Caribbean shear zone.

A NEW MODEL FOR THE BREAK-UP OF PANGEA

The data in the present study emphasize that the break-up of Pangea started during the Middle Triassic. The break-up started from the western termination of the Paleotethys Gulf. This first event corresponds to the opening of the Ladinian–Carnian Eastern Tethys ocean, located between the northeastern margin of the future Gondwana (in particular, the future Adria Mesoplate) and southeastern Laurasia (the future Europe Plate) (Fig. 11). According to two contrasting interpretations, this new oceanic branch was either the westward progradation of the Paleozoic Paleotethys, or a new oceanic basin. The latter hypothesis surmises that this basin was opening to the southwest of the Paleotethys, cutting off a narrow and elongated strip of the Gondwana–Africa margin and continuing westwards into the Pangea continent (see Fig. 2). The remnants of this ocean opening are the Triassic MORB ophiolites that crop out along a disjointed belt, which ranges from the Pontic Ranges, throughout the Hellenides and Dinarides, and ends in the Western Carpathian (Pieniny Mountains). The Pieniny Mountains zone can be considered the western tip of the prograding Eastern Tethys ocean basin. If we 'remove' the orogenic deformations of late Mesozoic and Tertiary age, the most probable geographic trend of this basin during the Triassic–Early Jurassic was probably ESE–WNW. This ocean began to close during the Early Jurassic, while the spreading was still continuing, and it was completely subducted by the Jurassic–Cretaceous boundary. When this ocean began to close, the Western Tethys began to open along a WSW–ENE trend, and continued its spreading until the Kimmeridgian–?Tithonian. The Western Tethys began to close in the Late Cretaceous and was completely closed only by the Eocene. Between the Kimmeridgian–?Tithonian and the Late Cretaceous, the Western Tethys was a fossilized ocean basin (see Appendix i), and Iberia, the Western Tethys and Adria constituted a single composite plate (Abbate *et al.* 1984).

The geodynamic releasing from the southern and northern mega-plate would probably run, to the north, along the paleo-Pyrenean line, if it extended eastwards, as far as joining the northern border of the active Dinaride orogenic belt (see Stephan *et al.* 1990 and Schettino & Scotese 2002).

These two ocean basins (Western and Eastern Tethys) coexisted for approximately 20 Ma (from Bathonian–Bajocian to Kimmeridgian–?Tithonian). The possible link between the north-western termination of the Eastern Tethys and the north-eastern end of the Western Tethys is still a matter of debate. A possible interpretation (e.g. Abbate *et al.* 1980; Dal Piaz *et al.* 1995) surmises that the two oceans were connected and, possibly, that they both terminated against a large transcurrent fault line that could have been the eastwards continuation of the paleo-Pyrenean line.

In addition, the Central Atlantic Ocean began to open during the Bajocian–Bathonian interval, at the same time as the Western Tethys but, unlike the former, it has been active up to today (see Appendix i).

The age of the opening of the proto-Caribbean Ocean is a vexed question. The circum-Caribbean MORB ophiolitic rocks, both Late Jurassic and Early Cretaceous in age, if of Caribbean origin would suggest a Late Jurassic–Early Cretaceous opening of the Caribbean Tethys.

The conclusions of the present discussion point to a Pangea break-up begun during Triassic time in the Eastern Mediterranean, during the Middle Jurassic in the Western Tethys and Central Ocean, and during the Late Jurassic in the Caribbean Tethys. The break-up was never complete: in the westernmost Mediterranean the two supercontinents moved along a main wrench fault zone.

There are many doubts about the alleged existence of a Permo-Triassic ocean south of Turkey (the Neotethys; see Stampfli *et al.* 1999), the closure of which would have produced the Upper Cretaceous IAT of the Taurus Belt, Cyprus and Hatay-Baer Bassit. The only remnants of this ocean are the 'basaltic rocks' of the Mamonia mélange of Cyprus. These basalts, with a WPT and MORB signature, occur interbedded and covered by pelagic *Halobia* limestones (Dilek & Flower 2003). This is the typical situation of Triassic rifts characterized by an attenuated continental crust that did not necessarily evolve into an ocean basin. Moreover, this supposed ocean would have been fossilized and without any tectonic activity from the Triassic to the ?Late Cretaceous, in the paleo-East Mediterranean, a very active zone.

The proposed evolutionary model for the Pangea break-up can be subdivided into three main periods.

MIDDLE–LATE TRIASSIC

The Middle–Late Triassic period (from the Anisian) constitutes the first main rifting stage, which produced in Pangea a frame of extension structures distributed, as rift basins, mainly along the alignments in which, successively, all Mesozoic Tertiary ocean basins were generated (Fig. 12A). The opening of the Eastern Tethys began afterward, during the Ladinian–Carnian, and the opening was located between the western limit of the paleo-Tethysian Gulf and the Western Carpathians.

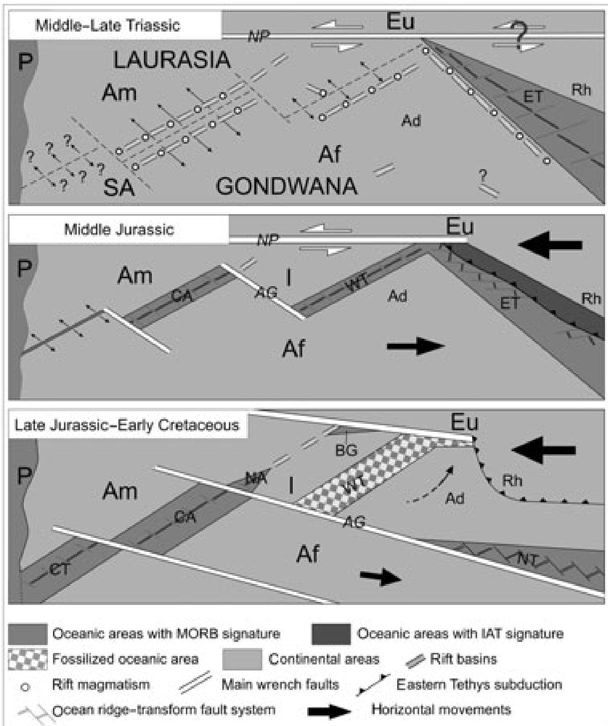

Fig. 12 Very schematic reconstruction of the Pangea break-up from the Middle–Late Triassic to the Late Jurassic–Early Cretaceous. Af, Africa with Adria Promontory (Ad); AG, Azores–Gibraltar Fault; Am, America; BG, Bay of Biscay; CA, Central Atlantic; CB, Caribbean Tethys; ET, Eastern Tethys; EU, Europe with Rhodope Massif (Rh); I, Iberia; NA, North Atlantic (Tagus Basin); NP, North Pirenean Fault; NT, Neotethys; P, Pacific Ocean; WT, Western Tethys.

EARLY MIDDLE JURASSIC

During the Bajocian–Bathonian, extensive sea floor spreading affected the Middle Triassic rift basins system in the Western Tethys and the Central Atlantic (Fig. 12B).

In the Eastern Tethys, spreading continued throughout the early Late Jurassic. During the same time period, an intraoceanic subduction occurred, with the consequent birth of a supra-subduction basin. In the supra-subduction domain, an intraoceanic obduction of the new IAT lithosphere onto the trapped MORB crust also occurred, and produced the metamorphic amphibolite soles.

In the south peri-Caribbean zone, a rifting stage took place, which led to the Late Jurassic–Early Cretaceous opening of the paleo-Caribbean Tethys.

LATE JURASSIC–EARLY CRETACEOUS

During the Late Jurassic, the collisional stage began to affect the Eastern Tethys domain, along an east-dipping subduction zone, and, contemporaneously, the Neotethys Taurus Ocean began to open (Fig. 12C). In the Western Tethys domain, drifting ended, and the ocean became fossilized. During the same period, sea floor spreading was incipient to the west of Iberia (Tagus Basin), in the Bay of Biscay, and in the Caribbean, while the Central Atlantic was drifting.

During the Early Cretaceous, the so-called ‘Gulf of Mexico Ocean’ ended active opening.

CONCLUSIONS

1. There is no close time succession between Triassic rifting and sea floor spreading.
2. Triassic sea floor spreading occurred only in the eastern peri-equatorial margins of Pangea, and did not cause important break-ups of the megacontinent. They can be considered as only the products of the westward propagation of the Paleotethys or Neotethys.
3. The Triassic Eastern Tethys was composed of only one oceanic branch (Dinaric–Hellenic–Pontic) as proved by the Triassic ophiolites. The presence of a Triassic Neotethyan branch south of Turkey is hardly tenable.
4. The Pangea break-up was completed during the Early Middle Jurassic in the peri-Mediterranean area, and in the Central Atlantic only during the Late Jurassic–Early Cretaceous in the Caribbean Tethys. This could be the time when the left-lateral strike-slip movement between Gondwana and Laurasia began.
5. The East Mediterranean Neotethys could have opened during the Early Cretaceous as a consequence of the continuing left-lateral movement between Gondwana and Laurasia, of the collision between Adria and Europe (Dinaric–Hellenic–Pontic orogenesis), and of the end of the drifting of the Western Tethys.
6. The separation of the two supercontinents was not realized through a continuous oceanic scar but, between the Western Tethys and the Central Atlantic Ocean, with a succession of small ocean basins, successively opened, going from east to west, and linked by east–west transcurrent fault structures. The short period in which the active oceanic areas between the two supercontinents were larger was probably between the Callovian and the Tithonian.

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APPENDIX I

During the preparation of this work, some ideas about the main general characteristics of the oceanic basins were discussed. In fact, it is immediately possible to distinguish at least three types of oceans: (i) long-lived oceans (e.g. the Central Atlantic); (ii) short-lived oceans (e.g. the Eastern Tethys and Caribbean Tethys); and (iii) fossilized oceans (the Western Tethys, the Gulf of Mexico? and the Ionian).

With special regard to the first two types, some considerations can be added:

1. The long-lived Atlantic Ocean, opened along the Paleozoic orogenic belts, seems to be the latest example of the several openings end-closures documented in the parallel orogenic chains accreted since the Middle Proterozoic along the eastern North America margin. The evolution history of these oceans is on the rank of, or bigger than, the Wilson cycle.
2. The short-lived oceans (e.g. the Eastern Tethys ophiolitic basin), on the contrary, develop during infra-Wilsonian cycles. If we consider in addition the more recent Indian Ocean, Red Sea, etc., a continuous process that detaches terrains from the passive margin of Gondwana and accretes them to the active margins of Eurasia can be hypothesized. This process seems active from the Permo-Triassic until today

It is difficult for us now to hazard logical explanations about the existence of some type of memoir located in the interior of the Earth, which would be responsible for these cyclic phenomena. If such memoirs exist, they could be located in the mantle or in the lithosphere. If the lithosphere continuously moves westwards (see Richard *et al.* 1991) in respect to the mantle, as documented by the movements of the hot spot frame, these memoirs are probably located in the lithosphere.

For case 1, a possible mechanism could be found, for the linkage between the orogenic belt and the new ocean opening. It could be caused by the asthenospheric rebound during the isostatic rise of the orogenic roots. According to Uyeda and Kanamori (1979), in the orogenic belts, the west-dipping subductions go to great depth, dragging down the orogenic roots. This fact could generate a very exceptional swallowing up of the asthenosphere. The rise of the asthenosphere could cause its partial melting and mantle denudation, triggering the sea floor spreading.

For case 2, it is very difficult to find a reasonable long-lived slicing mechanism responsible for the continuous transfer of Gondwanian terranes against the Laurasian southern border.